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Modelling climate change: an assessment of sea-ice and surface albedo feedbacks

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Abstract

The net strength of the sea-ice and surface albedo feedbacks in simulations of climate change has been estimated for several general circulation models (GCMs) but by a variety of different methods. Here, several methods are applied to quantify this feedback in a single CO_2 doubling experiment with the UK Meteorological Office GCM. Different methods give different values and are shown to be estimating different quantities. An unbiassed comparison of models is made using consistent methods, and an attempt is made to explain the differences in the models' sensitivities in terms of their control climates and their parametrizations. Comparing the standard experiment with the UK Meteorological Office model to a parallel experiment in which sea-ice extents were prescribed allows examination of the feedback as it actually operates in the model. Statistical methods are also used to examine the relationship between surface albedo changes, which are given directly by the surface heat balance and albedo parametrization, and planetary albedo changes, which incorporate the effects of clouds and determine the net feedback.

1. Introduction

It has long been realised that changes in surface albedo with temperature because of changes in snow and ice cover might provide a significant positive feedback on climate change (e.g. Croll, 1875). The feedback due to sea-ice changes on warming induced by increased atmospheric CO_2 levels has been a particular subject of discussion (e.g. Spelman and Manabe, 1984). Those general circulation model (GCM) experiments which allow the ocean to respond thermally can include this effect, and attempts have been made to evaluate its strength, considered as a simple feedback factor on global-mean time-mean surface temperature (e.g. Dickinson, 1986). However, this is known to vary considerably between models. It depends not only on the sea-ice extents in the control experiment, which are directly comparable with observations, but also on the parametrizations used. The latter are rather less easy to verify, yet they have a major effect on how readily the ice cover changes and also on how much of an albedo change a given change in ice cover produces.

The spectrally integrated albedo of sea-ice is dependent on the thickness, state and age of the ice itself and of any snow cover, on the solar zenith angle and, since cloud absorption is wavelength dependent, on cloud cover (e.g. Shine and Henderson-Sellers, 1985). An observation of sea-ice or snow albedo is made for a particular surface state at a single point, and is not easily compared to an albedo from a GCM, usually calculated from surface temperature meaned over a much larger area. (Strokina (1980) directly related temperature and albedo observations using data whose spatial scale was more appropriate for GCM modelling, but the timescales were much longer, and the temperature used was air temperature rather than surface temperature.) Even in a model which does calculate a prognostic snow cover over sea-ice, model biasses may mean that using this as the input to a physically reasonable parametrization would actually give less realistic results than a cruder scheme. There is therefore some difficulty in verifying parametrizations and modellers may "tune" their albedo formulations to give a good control simulation. For example, Manabe and Stouffer (1980) used unrealistically low albedos (with the resulting increased heat input into high latitudes tending to make up for their model's complete neglect of ocean transports) while Washington and Meehl (1983) used unrealistically high albedos. Shine and Henderson-Sellers (1985) have warned that such tuning is not necessarily advisable and even in an uncoupled sea-ice model may conceal serious errors in other aspects of the model formulation. Washington and Meehl (1986) have shown there may be considerable effects on model climate sensitivity.

It is therefore desirable not only to estimate the strength of the feedbacks associated with sea-ice and snow cover but also their interactions with other feedback processes, and explain these as far as possible in terms of model formulation and control climate. However, the various studies to date that have tried to quantify the strength of albedo feedback in a particular GCM have used different methods, and it is not clear whether the resulting estimates are truly comparable. Also, since the albedo changes are so restricted geographically, it is of interest how the warming which can be attributed to these changes is distributed. Much of the discussion of GCM climate change experiments has been in terms of global annual-mean surface temperature, but it is the geographical, seasonal and synoptic distribution of temperature change that is of economic and social importance.

This paper addresses some of these issues, using as an example the CO_2 doubling GCM experiment of Wilson and Mitchell (1987) (hereafter WM87). It aims to

(a) Analyze the mechanisms producing the sea-ice albedo feedback in this model

(b) Compare the strength of the sea-ice and snow albedo feedbacks in this model as estimated by various methods

(c) Compare the strength of these feedbacks consistently between the various models

An experiment parallel to that of WM87 in which sea-ice feedback was largely removed by prescribing sea-ice extents is discussed. Comparing the two experiments indicates the effect of the sea-ice feedback in the experiment with the full model. Section 2 describes the treatment of the ocean and sea-ice in the model used. Section 3 describes the distribution of the warming, the effect on low latitudes of the albedo feedback, and the mechanisms producing the albedo changes. Section 4 compares the strength of the sea-ice and surface albedo feedbacks between different models and between different methods of estimating it. Section 5 then provides a concluding discussion.

2. Model description

The atmospheric model used was a version of the U.K. Meteorological Office (UKMO) 11-level GCM. It has a penetrative convection parametrization, stability-dependent boundary-layer mixing, model-generated cloud amounts, prescribed cloud optical properties and the full seasonal and diurnal cycles (WM87). It was coupled to a 50 m "slab" ocean. The evolution of mixed layer temperature T at each point is given by:

$$\rho h c_p \frac{dT}{dt} = S + C \qquad (1)$$

where ρ and c_p are the density and specific heat capacity of water, h is the mixed layer depth, S is the total downward heat flux at the sea surface simulated by the atmospheric model and C is the oceanic heat convergence. A calibration integration run with the atmospheric model in which sea surface temperature (SST) and sea-ice extents were prescribed from climatology, so that dT/dt was specified and S simulated, allowed C to be found; then, in the integrations with the interactive ocean, this C was prescribed and used with the simulated S to find dT/dt. In practice C not only represents the convergence of heat by oceanic processes, but also compensates for deficiencies in the atmospheric model, and for the actual ocean mixed layer not being 50 m deep. It makes it possible to keep the SST distribution, and hence the atmospheric circulation, in the control simulation as near as possible to observations without affecting the model's freedom to respond thermally or introducing unrealistic parametrizations which might tend to affect the model's sensitivity. This should minimize the distortion of the model's response to climate change forcing (Mitchell et al, 1987). Monthly mean values of C were calculated and linear interpolation used between them, allowing detailed representation of the seasonal cycle while smoothing out synoptic time-scale variations in the forcing data.

Changes of state, and of ice thickness at ice-covered points, were also calculated every 5 days. The slab model formed sea-ice where a net cooling would have taken the slab temperature below 271.2 K, the freezing point of sea-water. Thereafter surface temperature was calculated every timestep (20 minutes), as at land surface points, but with the addition of a diffusive heat flux through the ice and a constraint that it could not be heated above freezing point. Any heating tending to do so was used to reduce the ice thickness at the end of the 5-day period. The snow-depth over sea-ice was calculated as a prognostic variable and used in the calculation of the thermal balance, both its latent heat of fusion and its effect on the diffusion of heat through the ice being allowed for. Values of the thermal diffusivities used were 2.166 $Wm^{-1}K^{-1}$ for ice and 0.3299 $Wm^{-1}K^{-1}$ for snow, as used by Semtner (1976), with densities of 900 kgm^{-3} for ice and 300 kgm^{-3} for snow. In the diffusion calculations the temperature at the bottom of the ice was taken to be 271.2 K, consistent with the heat balance, but a minimum ice thickness of 1 cm was added to that given by the heat balance calculations, so as to prevent unreasonably large fluxes through very thin ice.

The use of the prescribed heat convergence had been expected to give a good simulation of the seasonal evolution of the sea-ice edge in both

hemispheres. However, sea-ice was found to be over-extensive in development runs. This was because the calibration integration with prescribed SSTs and ice extents contained imbalances in its heat budget, associated with the prescription of sea-ice thicknesses of 2 m. These imbalances were not taken into account by the heat convergence, which was set to zero for periods when a point was ice-covered, and were exacerbated by albedo feedback. To counteract this the ice albedos were reduced from the values used in the calibration integration. In the model sea-ice albedo is linearly dependent on simulated surface temperature over a range below the melting point. This gives a simple representation of the effect of rising grid-box mean temperature in increasing the fractional coverage by melt-ponds and leads, and aging of any snow cover. Values of sea-ice albedo of 0.7 at and below 261.2 K, falling to 0.4 at 271.2 K were found to give satisfactory results with the "slab" ocean. These values are similar to those derived by Robock (1980) from detailed analysis of observations. In addition, when a non-zero heat convergence was specified for a point that happened to be ice-covered, it was used to melt ice if positive but distributed over the ice-free sea points of the hemisphere if negative. This conserved energy while again providing a local heating where it was needed, giving a very realistic control simulation of ice extents, as shown by WM87. The same parametrizations were used in all the integrations to be described. The use of the heat convergence at sea-ice points only had a significant effect where sea-ice was simulated but was not present in the climatology. Thus it is unlikely to have affected the model's response to a climate *warming*, in which sea-ice would be less extensive than in the control climate and in the climatology. The reduction in sea-ice albedos may have altered the strength of the sea-ice albedo feedback, but the values used remain within the range of observational uncertainty.

All the figures above refer to the model's "clear-sky" surface albedo: it should be noted that in the presence of cloud the actual surface albedo used by the radiation calculations is decreased to allow for multiple reflections between clouds and the surface, an effect of considerable importance over highly reflective surfaces (Slingo, 1985). This effect is also allowed for, in one way or another, in the other models to be discussed later.

Experiments. WM87 reported a CO_2-doubling experiment with the full atmospheric and "slab" model described above. We also shall show various results from it, all taken from 10 year means, from the 21st to the 30th year of the control integration and the 36th to the 45th year of the doubled CO_2 integration. This experiment showed a large climate sensitivity and large reductions in sea-ice cover.

In order to study the overall strength and the geographical distribution of the sea-ice feedback, a parallel experiment was set up with sea-ice extents prescribed for both the control and doubled CO_2 integrations. For every 5-day period they were taken from the same climatology originally used for the calibration integration. While open sea was prescribed, SST was calculated from the heat balance, even if it went below freezing point. While ice was prescribed, the normal heat transfer and ice depth calculations were carried out, but the ice thickness was not allowed below the minimum of 1 cm. Energy was conserved in this case by storing the extra heat and using it to increase the SST at the start of the next open-sea period. In general, of course, the model would not simulate an SST of precisely 271.2 K at the end of the climatological open-sea period, nor an ice depth of precisely 1 cm at the end of the climatological ice-covered period. To conserve energy with the least

distortion of the physics, the value actually simulated was therefore stored at the time of the prescribed change of state, to be used as the initial SST or ice depth at the reverse change of state. Extra heating tending to melt ice at permanent sea-ice points after the minimum ice thickness of 1 cm was reached was still lost to the system, but even in the doubled CO_2 integration this was small, only 0.17 Wm^{-2} in the global mean. Thus conditions during the frozen period and during the open-sea period at a seasonal sea-ice point were comparatively independent, though it was still possible for some heat to be transferred from the ice-covered period to the ice-free period. The ice bottom temperature was still taken as 271.2 K for the diffusion calculations.

A control integration with this alteration to the treatment of sea-ice was run for 7 years starting from the end of the 19th year of the control integration of WM87 and a doubled CO_2 integration for 17 years starting from the end of the 35th year of their doubled CO_2 integration. Both integrations appeared to have reached equilibrium, as shown by the evolution of the annual global mean surface temperature (Fig 1). Results shown hereafter from these integrations are means of the last 5 years of each.

3. The response to sea-ice changes in the UKMO model

In the standard CO_2 doubling experiment (WM87) global-mean annual-mean surface temperature increased by 5.2 K. In the prescribed sea-ice experiment this was reduced to 4.2 K. The global-mean annual-mean change in many other quantities was also reduced by a similar ratio, in accord with the usual observation that an AGCM's response can be approximately scaled by the mean temperature increase (Table 1). However, the geographical and seasonal distribution of change is also of importance.

3.1 Warming at high latitudes and seasonal heat storage. The albedo-temperature feedback in the experiment of WM87 might be expected to be most active in midsummer, when insolation is greatest, and to have no effect in the polar night. In fact, by far the greatest change in surface temperature in areas of reduced annual-mean sea-ice cover is in winter (compare Figs 2 and 3). This effect is discussed in detail by Manabe and Stouffer (1980). Although much more heat is absorbed in summer, it generally goes either into the slab ocean with its high heat capacity, and is mostly stored until autumn, or into melting ice whose surface was already at melting point. In either case it increases the total (sensible and latent) heat content of the mixed layer while giving little or no immediate change in surface temperature. However, this increased heat content considerably delays re-freezing later in the year, and the consequent large fall in surface temperature following the establishment of ice cover. Also, even after the surface has frozen over, there is a slightly larger diffusive heat flux through the ice from the water below, since the ice, having formed later, remains slightly thinner throughout its existence. The effect of this extra heat is increased by the small heat capacity of the frozen surface and the strong surface inversion present at high latitudes in winter, which confines its effect to near the surface. The latter is a more general effect well-known to give poleward amplification of AGCM surface response to climate change, and is not dependent on surface type or the presence of a seasonal cycle. In short, in the warmer climate sea-ice points absorb more heat in summer when the effective surface heat capacity is large, and release it in autumn and winter when the heat capacity is small.

In the experiment with prescribed sea-ice limits, this high latitude winter amplification of the warming was much reduced (Fig 2), because of the absence of the storage effect. The part of the sea-ice albedo feedback due to surface temperature changes remains, but is likely to be small because most insolation is received in local summer when the ice surface is generally already near melting point. Robock (1983) reported that this "snow/ice-meltwater feedback" was weak in his energy balance model. The thinner ice expected in the doubled CO_2 integration would still allow a greater upward heat flux through the ice at the same surface temperature. However, as indicated at the end of Section 2, the method used to prescribe the sea-ice limits prevented heat being passed from the open-sea period to the frozen period at seasonal sea-ice points, though it was possible for the reverse to happen. Thus either the surface warming and ice thickness must adjust to give an average zero heat flux through the ice over the frozen season, or there will be a net downward flux which will warm the ocean in the open-sea season.

3.2 Warming at low latitudes, and changes in atmospheric transports. The extent to which the warming due to reduced sea-ice extents penetrates to lower latitudes is also of interest. This can be estimated by comparing the experiment with prescribed sea-ice extents to the standard experiment of WM87. In the tropics the warming is reduced by about half a degree to 3-4 K, and in mid-latitudes by about 1 K to 4-6 K. The reduced warming at low latitudes is due to changes in the zonal-mean atmospheric transports of energy out of low latitudes, and the accompanying changes in the radiative heat balance (Table 2). In the experiment with prescribed sea-ice extents there is a greater increase on doubling CO_2 in the net flux out of low latitudes due to atmospheric processes than in the experiment of WM87. This is as would be expected, since the absence of the high latitude warming due to sea-ice feedback would tend to increase the meridional temperature gradients.

In the standard experiment of WM87 there is little change in the net export of energy from this region, with more than 90 % of the radiative forcing being compensated by local radiative changes. Some of the components of the net export do change substantially. For example, all the latent heat transports increase in magnitude as the saturation specific humidity increases, and the fluxes at levels around the tropopause increase in magnitude as the tropopause moves upwards (not shown). However, the contributions of all these changes to the net transport tend to cancel. Similar effects were noted by Manabe and Wetherald (1975; 1980) in the GFDL model, Rind (1987) in the GISS model and Held and Hoskins (1985) in a simple GCM. The latter remark that "The factors that control....the....degree of compensation....remain obscure to us" but suggest a possible dependence on convection and boundary layer schemes. One would also expect a dependence on the distribution of feedbacks at different latitudes. If these match, largely by chance, little change in transports would be needed to keep the north-south temperature gradients in balance with the model dynamics. The vertically integrated temperature gradient hardly changes in the experiment of WM87 (not shown) as the large surface warming at high latitudes is confined near the surface, and the small surface warming at low latitudes is magnified aloft by moist convective processes.

3.3 Surface and planetary albedos, and the effects of clouds. The strength of the sea-ice albedo-temperature feedback depends not on the size of the surface albedo changes themselves, but on the corresponding changes in the albedo at the top of the atmosphere. These depend on the other radiatively important quantities, especially cloud, in the control climate, and their general tendency to respond to climate change, and their local changes resulting directly from sea-ice cover changes themselves. These interactions have been discussed by Barry et al. (1984) using satellite observations and a radiative-convective model, but do not seem to have been analyzed for any GCM.

Fig 4 shows the relationship between the changes in the two types of albedo in the two CO_2 doubling experiments reported here. Each point on the scatterplot represents the change in surface and planetary albedo, meaned from December to February, at one sea-ice grid-point. There is a correlation in each case, but they are of opposite sign. A similar relationship holds in the other seasons. In the standard experiment, the dominant changes in surface albedo are due to open sea replacing ice, and these changes naturally feed through to the planetary albedo. There is also a general decrease in planetary albedo due to the reduction in cloud cover described by WM87, and to increased solar absorption by water vapour and the CO_2 itself. However, in the experiment with prescribed sea-ice extents, changes in the clear-sky surface albedo are

restricted to those given by changes in the surface temperature of sea-ice only. These are small in summer when most insolation is received. Hence changes in the effective surface albedo, taking into account multiple reflection from clouds, depend more on the general reduction in cloud cover. This gives an increase in effective surface albedo while also directly reducing planetary albedo. Given this uninstructive relationship, no further analysis of the changes of the different types of albedo will be presented for the experiment with prescribed sea-ice.

Table 3, however, presents figures quantifying such relationships statistically for the standard experiment. Linear regression is applied to the fields of change in the different types of albedo over sea-ice points. Thus, taking for example the regression of actual planetary albedo on actual surface albedo, the quantity *a* indicates that a given change in surface albedo typically causes a change in planetary albedo only 0.178 times as large, and *b* that planetary albedo is reduced on average by 3.26 % independently of changes in surface albedo. The relationship between the changes in the clear-sky albedos at the top and bottom of the atmosphere is much stronger than those relationships where cloud has an effect. Only a quarter of the spatial variance of changes in planetary albedo can be attributed to changes in surface albedo, the remainder presumably being due to changes in cloud. The 3.3 % reduction in planetary albedo which is independent of surface changes is largely a consequence of the general reduction in cloud reported by WM87. These facts are no doubt related to the high shielding of sea-ice by cloud (a mean cloud cover of 78 % in the control integrations of both experiments). In addition, the global values prescribed for the albedos of clouds (20 % for high cloud, 60 % for all other clouds) may be unrealistically large for polar clouds.

4. Comparison of estimates of the net feedback from sea-ice and surface albedo

4.1 Introduction. The importance of surface albedo feedback to AGCM climate change is generally acknowledged and several recent studies attempt to quantify it. The resulting estimates are summarized in Table 4, in terms of the "feedback parameter" used, for example, by Dickinson (1986). The use of this quantity assumes that the radiative response of the climate system to temperature change is proportional to the change in temperature (global annual mean surface temperature is normally used). The constant of proportionality is the total feedback parameter λ. Thus in particular, when the climate system has reached equilibrium with a radiative perturbation ΔF by changing its temperature by ΔT_{EQ} from its previous equilibrium,

$$\lambda \, \Delta T_{EQ} = - \Delta F \qquad (2)$$

and the magnitude of λ is an inverse measure of the overall sensitivity of the climate system. If separate processes give independent radiative responses the total feedback parameter can be expressed as the sum of individual feedback parameters for each process. For example, one might analyze λ:

$$(\lambda_{LW}+\lambda_{SA}+\lambda_{CS}+\lambda_{GS}) \, \Delta T_{EQ} = - \Delta F \qquad (3)$$

where the subscripts LW, SA, CS and GS refer respectively to all longwave radiative processes combined, to surface albedo processes, to the effects of clouds in reflecting insolation and to the effects of gaseous absorption of insolation. In such an analysis it is thus λ_{SA} that quantifies the strength of surface albedo feedbacks. Given an estimate ΔF_{SA} of the actual change in net radiative heating of the climate system due to surface albedo changes, λ_{SA} can be found:

$$\lambda_{SA} = \frac{\Delta F_{SA}}{\Delta T_{EQ}} \qquad (4)$$

with similar expressions for each of the other processes separately represented in Eq 3 or an analogous equation.

The sign convention adopted here means that positive feedbacks have a positive feedback parameter. Thus the total feedback parameter is negative for a stable system, which responds so as to cancel rather than add to any perturbation. A larger positive feedback parameter for, say, surface albedo processes thus gives a greater reduction in the magnitude of the total feedback parameter and a greater sensitivity.

4.2 Methods used to estimate the net feedback The methods used to derive the values in Table 4 vary as widely as the estimates they obtain.

Method 1, comparing the results of running the full AGCM (a) with interactive sea-ice and (b) with changes in sea-ice extents prevented, has already been described. It is physically the most direct of the methods in that it literally answers the question of how the response of the AGCM would differ with and without changes in sea-ice cover. The strength of the original forcing ΔF is the same in both cases, so the different ΔT_{EQ}s give different λs for the

two experiments. In the experiments with the UKMO model described above, ΔF from the doubling of CO_2 levels is about 4 Wm^{-2} and the two ΔT_{EQ}s are 5.2 K and 4.2 K. This gives λs of -0.95 and -0.77 $Wm^{-2}K^{-1}$, and their difference 0.18 $Wm^{-2}K^{-1}$ is the feedback parameter measuring the strength of the feedbacks from the sea-ice processes omitted in the second experiment.

Method 2a separates the changes in the solar radiation reflected to space due to surface albedo changes from those due to changes in cloud cover and water vapour absorption by assuming that surface albedo changes dominate in certain latitude bands and are negligible elsewhere. This allows a calculation of the mean radiative flux ΔF_{SA} associated with the feedback. ΔT_{EQ} is of course known, so that Eq 4 gives λ_{SA}. This method was introduced by Dickinson et al. (1987), who applied it to the National Center for Atmospheric Research (NCAR) model in the experiment of Washington and Meehl (1984). As well as its simplicity, this method has the great advantage of using only quantities diagnosed by the full AGCM, run without any distortion of its physics. However, it allows the calculation of feedback to be affected by changes in cloud correlated to the sea-ice changes. Dickinson (1986), in a brief but wide-ranging and valuable discussion of surface albedo feedback, also quotes values for various other studies without giving a source. They appear to be obtained by means similar to this method.

Method 2b is similar but uses correlation techniques to allow for the effect of clouds, as discussed in Section 4.3 below.

Method 3a is to run the radiation code of the model on time-mean model data. This can be done first using the time-mean values of the radiatively active quantities, temperatures, cloud amounts, concentrations of water vapour and other gaseous absorbers, and surface albedos, from one experiment. Then one of these fields, in this case surface albedo, is changed to the values from the other experiment, and the radiation code run again. The resulting change in the radiation fields thus provides an estimate of the change in the radiation fields in the full model due to the change in surface albedo, and is used as ΔF_{SA} in Eq 4. This was done by Roeckner et al. (1987) with the Hamburg University (HU) model, using separately meaned data for each month of the year. It has also been done with the GFDL model, as discussed below. This method should give a good diagnosis of the radiative effects of the time-mean changes. There is still the possibility of distortion due to the non-linear effect of variations about the mean, and the correlation in time of synoptic variations in different quantities.

Method 3b differs from Method 3a only in using un-meaned cloud amounts, as discussed in Section 4.5 below.

Method 3c is the same as Method 3a, except that rather than repeat the radiation calculations for each point and season it uses global-mean annual-mean changes, and also it uses a radiative-convective model (RCM) rather than just a radiation scheme. The RCM can be run to equilibrium with only one feedback process allowed to operate. This is intended to approximate the results of the impossible experiment of running the GCM with only that one feedback process. This method was used by Hansen et al (1984) to analyse experiments with the Goddard Institute for Space Studies (GISS) AGCM. While the global and annual meaning makes the analysis simpler than Method 3a, it obviously could introduce distortions.

It is important to note that these are not all merely different methods of finding the same quantity, differing only in accuracy or convenience: they do in fact include different sets of feedback processes. For example, Method 1 only finds the part of the surface albedo feedback due to sea-ice. Land surface albedo feedback is included only in so far as changes in snow cover, and so in land surface albedo, are forced thermally by changes in sea-ice extents. The other authors cited above applied their methods to the total surface albedo feedback. Although no authors to date seem to have quantified sea-ice and snow albedo feedbacks separately, they correspond to distinct physical processes and different model parametrizations. Indeed, Washington and Meehl (1984) indicate that in parts of the Northern hemisphere in winter their model gives increases in snow extent on increasing CO_2 and so at least locally a negative snow albedo feedback.

Another difference is that, except for Methods 1 and 3c, all of them are concerned with radiative quantities only, deriving a feedback parameter from Eq 4. Thus they do not explicitly take account of the seasonal storage effect, although it does have an effect on the feedback parameters calculated. To see this, let us consider for simplicity a hypothetical situation in which the seasonal storage effect increased ΔT_{EQ} as described in Section 3.1, but without affecting the radiative response at all. Then (Eq 2) the magnitude of the value of the overall, negative, λ, would be reduced in inverse proportion and (Eq 3,4) so would the magnitudes of each of its components as measured by the radiatively-based methods, including the only quantity measuring the effects of sea-ice, λ_{SA}. Yet since the effect on the overall negative λ, which measures the actual sensitivity of the system, is to reduce it, it is as if an additional positive feedback, with its positive λ, had been added to the system, and Method 1 would recognize this.

Although the seasonal storage effect increases the mean surface temperature, it does so merely by transferring heat from a surface with a high heat capacity to one with a low heat capacity, rather than by increasing the heat content of the system, and probably has negligible impact outside certain restricted areas. So it could be argued that this process is not really a positive feedback, and those analyses are correct to omit it. However, this is logically inconsistent with the fact that their feedback analyses are all carried out in terms of the changes in the GCM's global-mean annual-mean surface temperature, which the seasonal storage effect does increase. Method 3c, using only the surface temperature changes from the RCM, in which there is no seasonal storage effect, is of course logically consistent, even if inaccurate.

Applying the different methods described above to one model, the UKMO model, gives estimates varying by a factor of 2, as shown in Table 4 (compare Methods 1, 2a, 3a, and 3b). When the feedback strength is estimated in the same way the UKMO model agrees quite well with the HU and GISS models, but not with the NCAR model. We shall now compare the response of the UKMO model with each of the other models in turn and try to account for the differences in the feedbacks estimated.

4.3 Comparison with the NCAR model Dickinson et al. (1987) observed that in the CO_2 doubling experiment of Washington and Meehl (1984) the NCAR model produced very marked changes in net solar absorption by the surface-atmosphere system at certain latitude belts in each season (their Fig 1) where sea-ice and snow cover

were greatly reduced. With the comparatively small overall cloud feedback and generally small cloud changes in this experiment, they felt that a reasonable estimate of the surface albedo feedback could be obtained by assuming that this process was dominant there and negligible elsewhere (Method 2a of Table 4). In fact, their estimate of the global-mean cloud feedback should be slightly smaller still, about 0.35 $Wm^{-2}K^{-1}$ rather than 0.4 $Wm^{-2}K^{-1}$. This is because they estimated the cloud contribution as a residual after assuming that all the changes in solar absorption were due to feedback processes, whereas in fact about 0.25 Wm^{-2} of the 4.65 Wm^{-2} total change in solar absorption would be produced by the original forcing, the doubling of CO_2 levels. If local changes in cloud and in surface albedo are correlated then cloud changes can still affect such an analysis in a model with interactive cloud cover, however small the overall cloud feedback. Indeed, Washington and Meehl (1984) draw attention to "a relative decrease of clouds near 50° S and an increase near 70° S. This is due to a shift of the baroclinic zone and associated storminess in the circumpolar trough, which move southward with the corresponding shift of the sea-ice margin. This movement of the cloud maximum contributes to the ice-albedo feedback". Thus Dickinson et al. (1987) include the local changes in cloud associated with and apparently forced by the changes in sea-ice extents in their sea-ice feedback.

Method 2a would seem less suitable for the UKMO model, in which, even over sea-ice, changes in planetary albedo seem to be due more to cloud changes than to surface albedo changes (Section 3.3). Zones of surface albedo feedback could not be identified at all in the way Dickinson et al. (1987) did so, using the actual reflected insolation at the top of the atmosphere. This is because in the UKMO model these zones merge into the mid-latitude zones where the model produces a strong cloud albedo feedback. However, the WM87 model diagnosed not only the actual radiative fluxes but also the ones that would have been produced in the absence of cloud. Zones of large change in the clear-sky outgoing solar radiation were generally clearly identifiable. The zones given in Table 5 are polewards of those of Dickinson et al. (1987), given in their Table 1 (by 15° of latitude for the Northern hemisphere in summer and about 8° of latitude otherwise), as would be expected given the excessive sea-ice cover in the NCAR model. This will tend to reduce both the total area of the zones and the mean insolation they receive, so giving smaller contributions to the global mean feedback. The total insolation received by the zones chosen for the UKMO model is in fact 19 % less than for the NCAR model.

Using this method the UKMO model gives a larger feedback in the Northern hemisphere than the NCAR model in all seasons except summer (Fig 5). The bulk of the NCAR model's feedback is in the Southern hemisphere, where the UKMO model's feedback is weaker, and the overall mean is less than half as strong in the UKMO model. It is still almost twice the size of the feedback estimated for the UKMO model by the other methods (Table 4, Method 2a).

This is much as one might expect given the contribution from cloud changes, which are excluded by the methods used in other studies. Using regression data similar to that presented in Section 3a we can allow for this. The known mean change in surface albedo and the slope of the regression line give an estimate of the mean change in planetary albedo that the surface albedo changes alone would have produced. The ratio of this to the actual mean change

in planetary albedo can then be used to reduce the estimated feedback. Explicitly:

$$f_R = \frac{a\,\Delta\alpha_S}{\Delta\alpha_P} f_O \qquad (5)$$

where f_R and f_O are the reduced and original estimates of the feedback respectively, $\Delta\alpha_S$ and $\Delta\alpha_P$ are the mean changes in surface and planetary albedo respectively, and a is as defined in Section 3.3. As Table 4 shows (compare Method 2b with Method 2a), this halves both the land and sea components of the surface albedo feedback. Separate regression coefficients and fractional albedo changes were used for land and sea-ice, for each season and for each hemisphere, but regression on annual mean fields for all the land and sea-ice of the globe together gave a reduction of similar magnitude.

It should be noted that the experiment of Washington and Meehl (1984) was probably not at equilibrium. They themselves did not indicate that they considered it was, confining themselves to a comment that "the major patterns of changes ... were very consistent from one year to the next". Convergence data shown by Schlesinger et al (1988) suggest that both the control and doubled CO_2 integrations were still cooling significantly. The definition of the feedback parameter assumes that the albedo responds to changes in the mean temperature immediately and in the same way throughout the model's evolution, so that there is no need to use equilibrium values to estimate it. However, in reality the sea-ice cover would be expected to lag the global-mean temperature, perhaps significantly. Thus the change in reflected insolation associated with the sea-ice changes would be expected to be underestimated, and so would the feedback parameter. Also, a colder simulation would be expected to give a greater albedo feedback, since the albedo changes will occur at lower latitudes where a retreat of sea-ice or snow cover over a given distance uncovers a greater area, and where the amount of insolation is greater at most times of year (e.g. Spelman and Manabe, 1984). This would still hold for albedo changes directly due not to surface albedo changes but to cloud changes associated with the retreat of sea-ice or snow. Thus the feedback that would have been produced if the model had started from its equilibrium climate would be expected to be larger again than the actual feedback in the experiment of Washington and Meehl (1984). This suggests a very great feedback compared to the other models referred to in Table 4. Much of this might be attributable to the coldness of this model. However, if the unrealistically cold equilibrium control climate of this model indicates deficiencies in its formulation, these might also tend to make its response unrealistic.

4.4 Comparison with the HU model As Table 4 shows, there is remarkably close agreement between the UKMO and HU model surface albedo feedbacks, estimated in the same way (Method 3a). Since the surface albedo parametrization and control integration's snow and sea-ice distributions have not yet been published, it is impossible to judge whether this is likely to be due to cancellation of differences. The two models certainly differ considerably in most other aspects of their formulation. One important difference, given the extent to which the cloud distribution appears in the above discussion, is the cloud scheme. The HU model's cloud scheme has a prognostic cloud liquid water variable and treats latent heat release and cloud formation consistently, unlike the diagnostic cloud scheme used in the UKMO model, where, at most model levels, cloud is assumed to form without latent heat release. Experiments with a similar cloud liquid water

scheme in the UKMO model (Smith, pers.comm.) suggest that this alone would not in fact greatly change the cloud *amounts* simulated or their response to climate change. However, cloud radiative properties are calculated interactively in the HU model, rather than being specified globally as in all the other models discussed, which have no physically-based cloud liquid water content. This is likely to mean that in the HU model the colder clouds that form over sea-ice and snow are likely to be optically thinner, and so less effective in obscuring the surface albedo changes.

4.5 Comparison with the GFDL model Several GFDL models have been used for climate change studies and Table 4 includes estimates of the surface albedo feedback parameter for three of them. The models used by Manabe and Stouffer (1980) and Wetherald and Manabe (1988) are fairly similar to the other models discussed, having realistic geography and the seasonal cycle. In other GFDL studies (Wetherald and Manabe, 1981; Spelman and Manabe, 1984) idealized geography was used, and in all but one experiment the seasonal cycle of insolation was omitted. Hence their results are not suitable for detailed comparison, but are included for completeness. It is perhaps curious that the two apparently more realistically formulated models are the furthest apart. It should be noted that, except in one of the experiments of Wetherald and Manabe (1988), cloud amounts were prescribed in these studies.

Spelman and Manabe (1984) ran their radiation code on time-mean data to give estimates of the direct radiative effects of the simulated surface albedo changes. The estimated feedback is thus directly comparable with those for the HU and UKMO models produced by the same method, Method 3a. Dickinson (1986) quotes values for two other GFDL experiments. Manabe and Stouffer (1980) showed a graph of the zonal-mean change in absorbed insolation, which showed considerable increases towards the poles over the small, nearly constant value at lower latitudes. Since cloud cover was prescribed in this experiment, surface albedo changes must be responsible. Combining this change in flux with the mean surface warming then gives an estimate for the feedback parameter of 0.3 $Wm^{-2}K^{-1}$. Wetherald and Manabe (1981) quoted only the global-mean change in absorbed insolation. Taking a suitable value for the global mean increase in gaseous absorption of insolation, which is not likely to be model-dependent, the remainder can again be attributed to surface albedo changes. We include the effect of CO_2 as well as of water vapour and obtain slightly lower values than Dickinson (1986), namely 0.35 $Wm^{-2}K^{-1}$ for the model with seasonal variation and 0.45 $Wm^{-2}K^{-1}$ for the model without. These two estimates should be comparable to that found for the UKMO model by Method 2b, which takes the appropriate change in insolation absorbed by the model as it runs, and corrects it for the cloud changes which the UKMO model allows. Wetherald and Manabe (1988) used Method 3b of Table 4, very similar to Method 3a. Since the response of the radiative fluxes to cloud amount changes is probably not close to linear, they feared there could be significant distortion introduced by the use of time-mean cloud amounts. To avoid this possibility, they ran their radiation code on seasonal mean values of temperature, water vapour and surface albedo for a large number of daily sets of cloud amounts. Reproducing this procedure with the UKMO model was not felt to be practicable, so that we have no precisely comparable figure for this model. However, we feel the difference between the results of Methods 3a and 3b is likely to be small compared to the differences between the models.

The values found are, then, around two to three times the size of those for the UKMO model, and vary by a factor of two. Various differences of

model formulation could account for this. Wetherald and Manabe (1981) demonstrated that the presence of the seasonal cycle reduced the climate sensitivity, since there is less snow and sea-ice in summer when there is most insolation. (Although most authors assume this to be the dominant effect of introducing or strengthening the seasonal cycle, Schlesinger et al., 1988, suggest that, especially in a warmer model, it could be outweighed by the tendency of the seasonal cycle to increase the control climate albedos, and so the possible albedo reductions, in winter.) Spelman and Manabe (1984) showed that the sensitivity of their model was dependent on the temperature, and hence areal sea-ice extents, in the control simulation. These two studies also differed in the parametrization of sea-ice albedo. Given the difficulties of comparing these two studies with ones using realistic geography and the seasonal cycle, we shall not discuss them further.

The experiment of Manabe and Stouffer (1980) parametrized the surface albedo of snow and sea-ice as a function of latitude rather than of temperature, with generally higher values than WM87. The experiments of Washington and Meehl (1986) with the NCAR model suggest that this should make it more difficult to melt ice, and so reduce the feedback from sea-ice retreat. However, as noted above, most of the feedback associated with sea-ice changes in the NCAR model may not be directly due to the sea-ice albedo changes but to the cloud changes stimulated by the sea-ice changes. This would mean that the strength of the feedback in the NCAR model would be largely determined by the distance the sea-ice retreats, rather than the direct radiative effects of its retreat. In the GFDL experiments with prescribed cloud the feedback associated with sea-ice changes must all be due to the surface albedo changes. These are affected in two ways by changing the albedo parametrization. Larger sea-ice albedos give larger local albedo changes where sea-ice melts, which tends to increase the feedback, but also make melting more difficult, which tends to reduce the feedback. Either effect could dominate in those cases where the feedback is due to the surface albedo changes themselves, so that we cannot make any assumption *a priori* about the effect of the different albedos in the models of Manabe and Stouffer and of WM87.

Manabe and Stouffer (1980) note that the sea-ice extents in their control simulation are more extensive than observed in the Northern hemisphere. This probably explains why the contribution from the Northern hemisphere is so much larger than in the UKMO model (0.26 $Wm^{-2}K^{-1}$ as opposed to 0.1 $Wm^{-2}K^{-1}$). The corresponding figures for the Southern hemisphere are 0.05 $Wm^{-2}K^{-1}$ and 0.06 $Wm^{-2}K^{-1}$ respectively. This apparent agreement is probably a result of cancellation between the effects of both the sea-ice extents and the cloud amounts in the GFDL model being rather less than observed. Factors such as differences in the surface albedo parametrization may also have an influence.

The other GFDL study with realistic geography and the seasonal cycle, Wetherald and Manabe (1988), is the furthest away from Manabe and Stouffer (1980) in terms of the net radiative feedbacks, with values of 0.55 and 0.59 $Wm^{-2}K^{-1}$. (The first value comes from the full model with interactive cloud, and the second from an experiment with cloud amounts prescribed from the control integration of the first experiment, but otherwise identical.) These values are the highest in Table 4 apart from the NCAR model. The surface albedo feedback would be increased by the fact that cloud cover is much too low in summer in the high latitudes of both hemispheres (their Fig 3). This may also be responsible for the models' excessive warmth in the southern hemisphere. Wetherald and Manabe (1988) do not document their model's albedo parametrization or sea-ice extents. Examination of its zonal mean planetary

albedo (their Fig 4) does not suggest any particular bias in either of these: possibly the effects of omitting all oceanic heat transport may be compensated by the extra insolation absorbed at middle and high latitudes because of the deficient cloud cover.

4.6 Comparison with the GISS model Hansen et al (1984) used Method 3c of Table 4, inserting into a radiative-convective model the global-mean annual-mean change in one radiatively active quantity and running the model to equilibrium while preventing other feedbacks from operating. They analysed the results to produce the gain g and feedback factor f, dimensionless quantities conventionally used in electronics. As they explain, the total g is defined as the fraction of the equilibrium temperature change due to all the feedbacks, and the total f as the ratio of the change with all the feedbacks to that with none. These total figures can then be split into separate figures associated with independent feedback processes, as for the feedback parameter λ discussed in Section 4.1. For each process, g and f are related by:

$$f = \frac{1}{(1-g)} \qquad (6)$$

with the result that the f do not combine linearly. Since Method 3c gives only surface temperature changes directly, this sort of analysis may seem more appropriate than one in terms of λ, which requires a radiative flux as well. However, we shall convert these values to λs for consistency with the rest of our analysis. The relationship between the λ and the g for the same process is:

$$\lambda = -g\,\lambda_{NF} = -g\,\frac{\Delta F}{\Delta T_{NF}} \qquad (7)$$

where λ_{NF} and ΔT_{NF} are respectively the sensitivity and the equilibrium surface temperature change with no feedbacks present. It should be noted that this surface temperature change "with no feedbacks", which is needed to calculate g and f or to convert them to λ, does not, strictly speaking, exist since all the RCM experiments include the "most basic" feedback, the temperature-dependence of the Planck function and of the gaseous absorptivities. In the RCM it is possible to carry out an experiment with only this one feedback, but in the real atmosphere or in the GCM other processes are just as fundamental. Thus the λ are closer to quantities with a real meaning in the GCM or in the real atmosphere.

A series of climate change experiments with the GISS model, differing in parametrizations or forcing, gave gains due to surface albedo processes of 0.09 to 0.12. Carrying out the same process with data from the experiment of WM87 gave a value of 0.07. Eq 7, assuming ΔF (at the tropopause) to be 4 Wm^{-2}, converts these to λs of 0.28-0.37 $Wm^{-2}K^{-1}$ for the GISS model and 0.30 $Wm^{-2}K^{-1}$ for the UKMO model (Table 4). The fact that the dimensionless quantities for the UKMO model are outside the range given by the GISS model, although the λ is not, is due to ΔT_{NF} in the RCM being rather less in the UKMO case (0.92 K) than the GISS case (1.3 K).

Thus the two models appear to give a fairly similar albedo feedback strength, especially comparing the "standard" GISS experiment with the UKMO

model. This might be expected given their general similarity in formulation and response to climate change forcing. In particular, the sea-ice extents in the control simulation of WM87 are similar to those in the GISS "standard" experiment's control, being slightly less than observed. One difference is that the GISS model uses a more complex sea-ice formulation, including allowance for partial cover of a grid-box by sea-ice, and a dependence of sea-ice albedo on snow depth and age (Hansen et al., 1983). The albedos during the melting season are probably most important, and there is little difference between the albedo of sea-ice at melting point in the UKMO model, 0.4, and the spectral-mean albedo of snow-free sea-ice in the GISS model, 0.45. The alternate GISS experiment, which has more extensive sea-ice in the control simulation, gives a stronger surface albedo feedback, as would be expected. There are also of course many other possibly relevant factors, most obviously cloud cover.

Surface albedo changes can have little effect on overall absorption of insolation if cloud cover is so great as to prevent much radiation reaching the surface. Dickinson (1986) stated that the ice albedo feedback in the experiment of Hansen et al (1984) "appears to have been strongly masked by high-latitude cloudiness". This would not, however, have reduced the value (0.3-0.4 $Wm^{-2}K^{-1}$) he quotes for its strength, if it is taken from the RCM experiments of Hansen et al. (1984), since these used global-mean data. It would tend only to reduce the feedback in the full AGCM, so that the estimate from the RCM experiments of the model's surface albedo feedback would be too great. This agrees with what was found with the UKMO model where using Method 3c rather than Method 3a increased the estimated surface albedo feedback by more than half (Table 4). This is almost entirely due to the geographical meaning. Calculations with global-mean data for each of the four seasons give results very close to one another and to calculations with global-mean annual-mean input data. Using a full RCM rather than using the radiation scheme purely diagnostically, directly analogous to Method 3a except for the averaging of the data, only increases λ from 0.28 $Wm^{-2}K^{-1}$ to 0.30 $Wm^{-2}K^{-1}$. Of course, these findings may not be true for other models. Indeed, the lack of a significant difference in the λs given by the full RCM and by the radiative-diagnostic model with globally meaned data in the UKMO case may well be due to a cancellation of differences. The RCM-simulated surface warming due to all processes is 6 K, rather larger than that given by the GCM, whereas if Method 3c did allow the various radiative feedbacks to operate realistically, the only difference to be expected would be a slightly *smaller* warming in the RCM because of the lack of the seasonal storage effect . Hansen et al did find good agreement between the RCM and the GCM for the surface warming, which they described as "circumstantial evidence that the procedure takes into account the essential radiative aspects".

Allowing for the distortion due to the use of global-mean data, the actual total radiative feedback directly due to surface albedo changes in the GISS model with the standard sea-ice parametrization may well be around 0.2 $Wm^{-2}K^{-1}$, as for the UKMO model. It should be noted that the high-latitude cloudiness of the GISS model is not necessarily excessive - Fig 26 of Hansen et al (1983) shows Arctic cloud cover in summer in the GISS model everywhere below 80 % and generally below 70 %, while the climatology of Orvig (1970) shows it to be almost everywhere over 80 % and generally over 90 %. Of course, cloud optical properties as well as nominal cloud cover are important, and a lower than observed cloud fraction can still be associated with excessive cloud forcing if the modelled clouds are optically thicker than in reality.

4.7 The feedbacks in the UKMO model We have seen that even those methods of quantifying albedo feedback which do produce similar values for the overall feedback strength when applied to the same model, the UKMO model, are not in fact estimating the same quantity. Two of the results can be considered to be seriously flawed. Method 2a includes the effects of local cloud changes, which in this model seem to be part of a global cloud response rather than a response to the local surface changes. Method 3c, at least in this model, seriously under-estimates the cloud shielding of surface albedo changes.

The other three methods can all be considered correct in their own ways, yet are still quite distinct and could potentially give three very different answers. Method 1 excludes most of the land surface albedo feedback and some of the sea-ice albedo feedback, but includes the seasonal storage effect. Methods 2b and 3a both aim to diagnose the radiative impact of the surface albedo changes free from the effects of the cloud feedback, but not in the same way. Method 3a should give an estimate of the effect of surface albedo changes completely separate from any cloud changes. Method 2b would leave in the effect of cloud changes actually correlated to surface albedo changes point-by-point, while removing the effects of, say, a shift in a cloud belt due to the surface albedo changes but not located at the same points. In fact, Methods 2b and 3a agree well, not only for the net feedback but also for its seasonal and geographical dependence (Fig 6, to be discussed below). This suggests that the cloud and surface albedo feedbacks are largely distinct in the UKMO model and that we may analyze them separately without introducing serious distortion. This is supported by the fact that the changes in radiative fluxes estimated by Method 3a are little affected by whether the other radiatively active quantities, including cloud amounts, are specified from the control experiment or from the doubled CO_2 experiment: not only are the global means similar but the spatial correlations are all over 97 %. However, the fact that Methods 2b and 3a also agree well overall with Method 1 must be due to chance cancellation of the many differences.

In Section 3.1, we used Method 1 to examine the seasonal and geographical distribution of the temperature changes, which are the final result both of the surface albedo feedbacks and, more importantly, of the seasonal storage effect. Methods 2b and 3a allow us to look at the distribution of the radiative changes that are the first stage in the albedo feedback process, as is done in Fig 6. Method 2a is also included for completeness and to show its fairly uniform over-estimation of the feedback. The Figure shows the contribution to the globally annually averaged surface albedo feedback parameter coming from each hemisphere and from each season. In the northern hemisphere land and sea are also separated. (The southern hemisphere feedback is all from sea-ice except for Method 2a, which contains a small contribution due to cloud changes over land.)

In the southern hemisphere the strength of the feedback appears to vary seasonally much as the insolation does. A similar pattern, but with both a larger mean and a larger amplitude of the seasonal cycle, is seen when we compare with the analysis of Dickinson et al. (1987), using the same method (Fig 5). The northern hemisphere seasonal values and those shown by Dickinson et al. are quite similar, given the differences in the zones included and in the general nature of the models' response. Dickinson et al. do not break down their northern hemisphere figures into land and sea, but Fig 19 of Washington and Meehl (1984) shows that at least in the solstitial seasons the feedback from the sea dominates. This is in contrast with the UKMO model in winter, where almost all of the feedback is from land. In summer both derive most of

their feedback from the sea, but from different latitudes: the entire Arctic Ocean is thawed in the UKMO model but the sea-ice retreat in the NCAR model is close to the Arctic circle. In both models the largest contribution in the northern hemisphere is from spring. In the UKMO model the spring contribution comes largely from land. It is associated with a tendency to excessive snow cover over the northern continents in spring in the control integration, when they warm up too slowly. However, this will be partly shielded by the excessive cloud cover also found, so that this contribution to the net surface albedo feedback may not be exaggerated. If the contribution from the northern continents in spring were, say, halved, the global annual mean surface albedo feedback given by Method 2b or Method 3a would be reduced by about a sixth (Fig 6).

5. Discussion

The strength and the realism of surface albedo feedbacks in GCMs has been the subject of some debate. We have shown that the estimated magnitude of the surface albedo feedback depends not only on the characteristics of the model used (domain, inclusion of the seasonal cycle, specification of surface albedos, etc) but also on the method used to estimate the feedback. Based on results from the UKMO model (Table 4) we find:

(1) Comparing CO_2-doubling experiments with the full GCM in which both interactive and prescribed sea-ice extents were used (Method 1) estimates the sea-ice feedback as 0.18 $Wm^{-2}K^{-1}$.

(2) Using changes in planetary albedo in selected regions, corrected by a regression technique to allow for cloud changes (Method 2b) and running the radiation code on gridpoint time mean data (Method 3a) give similar estimates of the surface albedo feedback, 0.16 to 0.18 $Wm^{-2}K^{-1}$.

(3) Using changes in planetary albedo in selected regions, without allowing for cloud effects (Method 2a, Dickinson et al, 1987) substantially over-estimates the surface albedo feedback.

(4) Running the radiation code on globally averaged time mean data (Method 3c, Hansen et al, 1984) also substantially over-estimates the surface albedo feedback.

Assuming where necessary that these findings are also true for the other models we can rank the models by their surface albedo feedback. The UKMO and HU models produce the smallest feedback. Allowing for the bias in Method 3c, the GISS model probably lies between these and the GFDL model (Manabe and Stouffer, 1980). The NCAR model gives the largest surface albedo feedback, though the value given in Table 4 is undoubtedly inflated by cloud changes.

The question of greatest interest is how great the actual feedback should be. The value in the experiment of Washington and Meehl (1984) is probably too large because of the coldness of their model at high latitudes, even if one accepts the local cloud changes as a legitimate part of the sea-ice feedback. We do not even know the sign of the response we should expect for cloud feedback. This suggests that we should as far as possible discuss albedo feedbacks and changes separately from cloud feedbacks and changes. The two effects in the NCAR model may, though, be so closely intertwined that they cannot be meaningfully analysed separately. As discussed above, the greater feedback given by the GFDL model of Manabe and Stouffer (1980) than by the HU, GISS and UKMO models may be associated with excessive sea-ice in high northern latitudes, suggesting that the smaller values are more accurate. However, at least for the comparison with the GISS and UKMO models, smaller cloud shielding by this GFDL model probably plays a part (as it certainly does for the model of Wetherald and Manabe, 1988). This is difficult to judge on the published model data, and in any case observational data for verification are not very helpful. Even if the GFDL model does have too little cloud, this may tend to compensate for the optical properties of the clouds in the GFDL, UKMO and GISS models, which have to prescribe global values for cloud optical properties which may be quite inappropriate for the thin stratus common in the Arctic. All this

suggests a value around the range 0.2-0.3 $Wm^{-2}K^{-1}$, but this must continue to be uncertain.

Future work should take care to define clearly just which feedback is being estimated, but further improvements to the GCMs used are needed. The parametrizations of cloud amounts and optical properties, as well as surface albedos, require improvement: they should both be firmly based on physical reality and also give realistic results when interacting with the other parametrizations in the control simulation. The models discussed in this paper either neglect changes in oceanic heat transport, or ignore the advection of heat by the oceans altogether. Furthermore, they all use extremely crude models of sea-ice, ignoring ice dynamics and simplifying many of the processes which contribute to ice formation and decay. Hence we must wait for fully coupled and physically realistic ocean-atmosphere-sea-ice models for a definitive estimate of the strength of sea-ice feedbacks.

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Table 1
Annual mean global-mean equilibrium changes due to CO_2 doubling

	With interactive sea-ice extents		With prescribed sea-ice extents		Ratio of changes
	Control	Change	Control	Change	
Surface temperature (K)	287.66	5.17	287.52	4.19	.810
Precipitation (mm/day)	3.02	.46	3.01	.37	.808
Surface albedo (%)	10.22	-.59	10.23	-.24	.41
Planetary albedo (%)	30.00	-2.86	30.02	-2.32	.811
Total cloud cover (%)	51.44	-3.83	51.51	-3.32	.868

Table 2
Annual mean changes meaned over 30° N to 30° S

	Radiative forcing* (Wm^{-2})	Atmospheric flux convergence (Wm^{-2})	Total forcing (Wm^{-2})	Surface warming (K)
Standard model (WM87)	4.79	-0.45	4.39†	3.95
Model with prescribed sea-ice extents	4.79	-0.72	4.07	3.40

*Estimated as the instantaneous change in net downward flux at the tropopause on doubling CO_2 levels in diagnostic runs of the radiation scheme of the type described in Section 4.
†Including a small contribution from the change in the oceanic heat convergence actually used, since this depends on the position of the sea-ice edge, as discussed in Section 2.

Table 3

Regression coefficients obtained by root-mean-square fitting of the regression line "$\Delta\alpha_1 = a\Delta\alpha_2 + b$" to changes in different types of albedo in the standard experiment in the annual mean over each sea-ice point.

α_1	α_2	"a"	"b" (%)	Correlation coefficient
Actual planetary albedo	Actual surface albedo	0.178	-3.26	0.51
Clear-sky planetary albedo	Clear-sky surface albedo	0.775	-0.24	0.9996
Actual planetary albedo	Clear-sky planetary albedo	0.196	-3.28	0.47
Actual surface albedo	Clear-sky surface albedo	0.905	0.37	0.98

All grid-boxes with sea-ice at any time in either integration or in the standard model climatology are included. Each is weighted by its area and the incoming insolation at the top of the atmosphere. Clear-sky radiation diagnostics were produced by the model for every point and timestep and give those radiative fluxes that would be calculated by the radiation scheme in the absence of cloud for an otherwise identical profile.

Table 4

Comparisons of different methods which have been used to quantify sea-ice and surface albedo feedbacks observed in AGCM climate change experiments

Method used to estimate feedback	Processes thus included	Processes thus excluded	Value found for the feedback parameter as defined in Section 4.1 ($Wm^{-2}K^{-1}$) in: UKMO model	Other model	Other model and source
(1) Comparing the effects of running the full AGCM with interactive sea-ice and with sea-ice extents prescribed from climatology	Changes in sea-ice extents, changes in cloud and snow cover thermally or radiatively forced by the former, seasonal heat storage effect.	Changes in land surface albedo not forced by changes in sea-ice extents, thermal changes in sea-ice albedo	0.18		
(2a) Assuming the change in planetary albedo to be due to surface albedo changes at certain latitudes and other processes elsewhere	Surface albedo changes locally, possibly significant co-located cloud changes.	Surface albedo elsewhere, non-radiative effects	0.31	0.7 0.3	NCAR, Dickinson et al (1987) GFDL, Manabe and Stouffer (1980)
As (2a) but for sea points only	Sea-ice albedo changes locally, possibly significant co-located cloud changes	Sea-ice albedo elsewhere land surface albedo, non-radiative effects	0.2		
(2b) As (2a) but adjusting to allow for cloud effects using the regression figures of Table 2.	Surface albedo changes locally	Surface albedo elsewhere, non-radiative effects	0.16		
As (2b) but for sea points only	Sea-ice albedo changes locally	Sea-ice albedo elsewhere, land surface albedo, non-radiative effects	0.09		

Method used to estimate feedback	Processes thus included	Processes thus excluded	Value found ($Wm^{-2}K^{-1}$) in: UKMO model	Other model	Other model and source
(3a) Running the radiation code alone diagnostically on time-mean data to isolate the effects of individual processes.	Surface albedo changes	Cloud changes, time variation, non-radiative effects.	0.17,0.18*	0.18+ 0.45+	HU, Roeckner et al (1987) GFDL, Spelman and Manabe (1984)
As (3a) but for sea points only	Sea-ice albedo changes	Land surface albedo, cloud, time variation, non-radiative effects	0.08		
(3b) As (3a) but run for a large number of instantaneous cloud distributions rather than the time-mean cloud distribution	Surface albedo changes	Cloud changes, time variation, non-radiative effects.		0.55, 0.59[1]	GFDL, Wetherald and Manabe (1988)
(3c) As (3a) but with seasonally and geographically meaned data, and using a radiative-convective model	Surface albedo changes	Variations in time and space	0.30	0.28,0.37[2]	GISS, Hansen et al (1984)

"Non-radiative effects" refers to the way in which seasonal storage of heat alters the mean warming, and so the value of the feedback parameter calculated from radiative quantities, independently of its radiative effects, as discussed in Section 4.2.

For clarity some of those experiments with the GFDL model with idealized geography, referred to in Section 4.5, are omitted since they are least comparable with the other experiments.

* Estimated using values for the other radiatively active quantities from control and doubled CO_2 integrations, respectively. Thus the smaller value is appropriate in the limit of a small change about the model's control climate.

\+ Estimated using values for other radiatively active quantities from the control integration.

1 Estimated from models with and without cloud changes permitted, respectively.

2 Extreme values, from various experiments differing in parametrization or forcing.

Table 5
Latitude belts where surface albedo changes were taken to be concentrated

DJF	MAM	JJA	SON
40°-60° N	50°-70° N	70°-90° N	50°-70° N
60°-80° S	60°-80° S	55°-75° S	55°-75° S

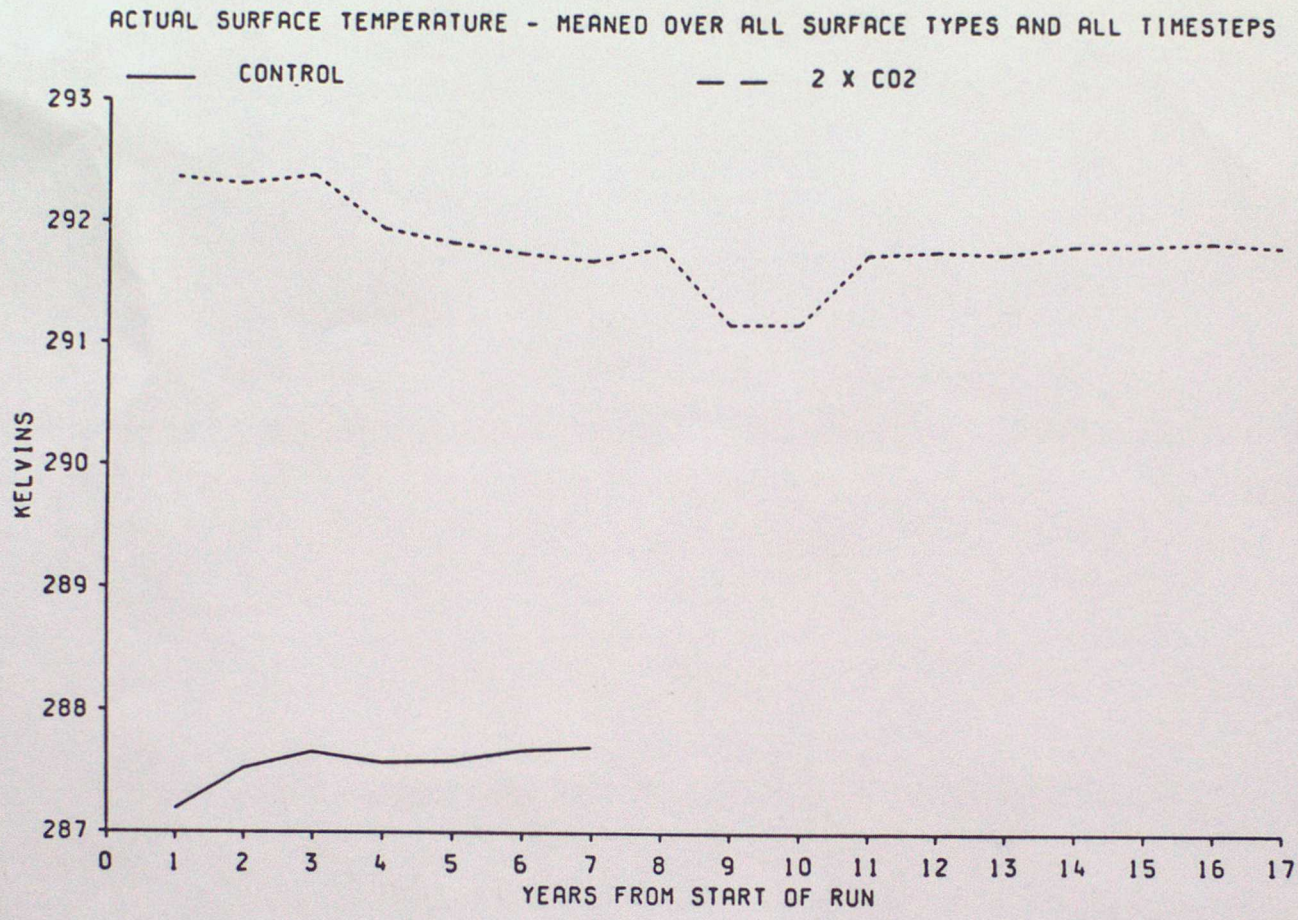


Figure 1. Evolution of global annual mean surface temperature in the control (solid line) and 2 X CO_2 (dashed line) integrations with prescribed sea-ice limits.

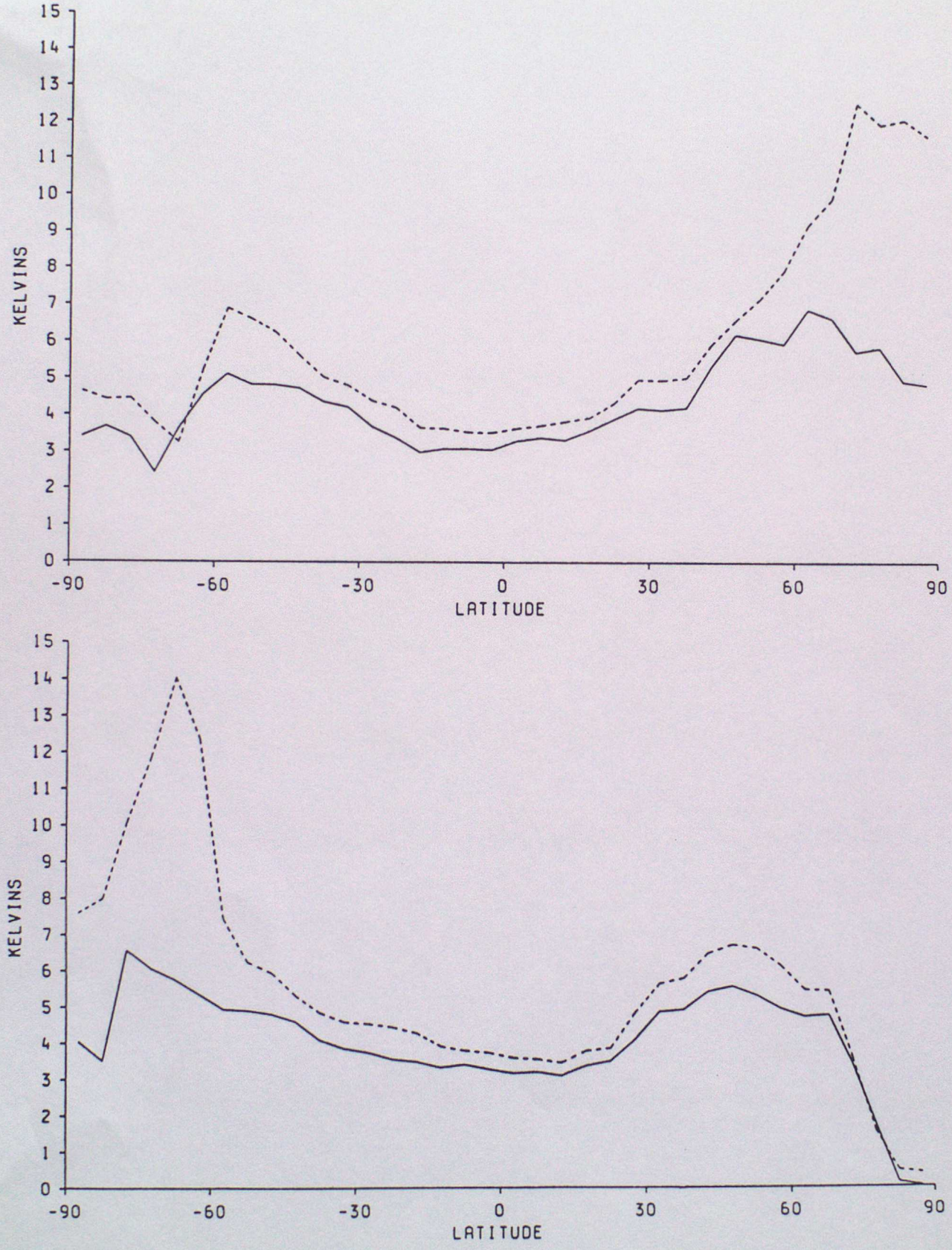
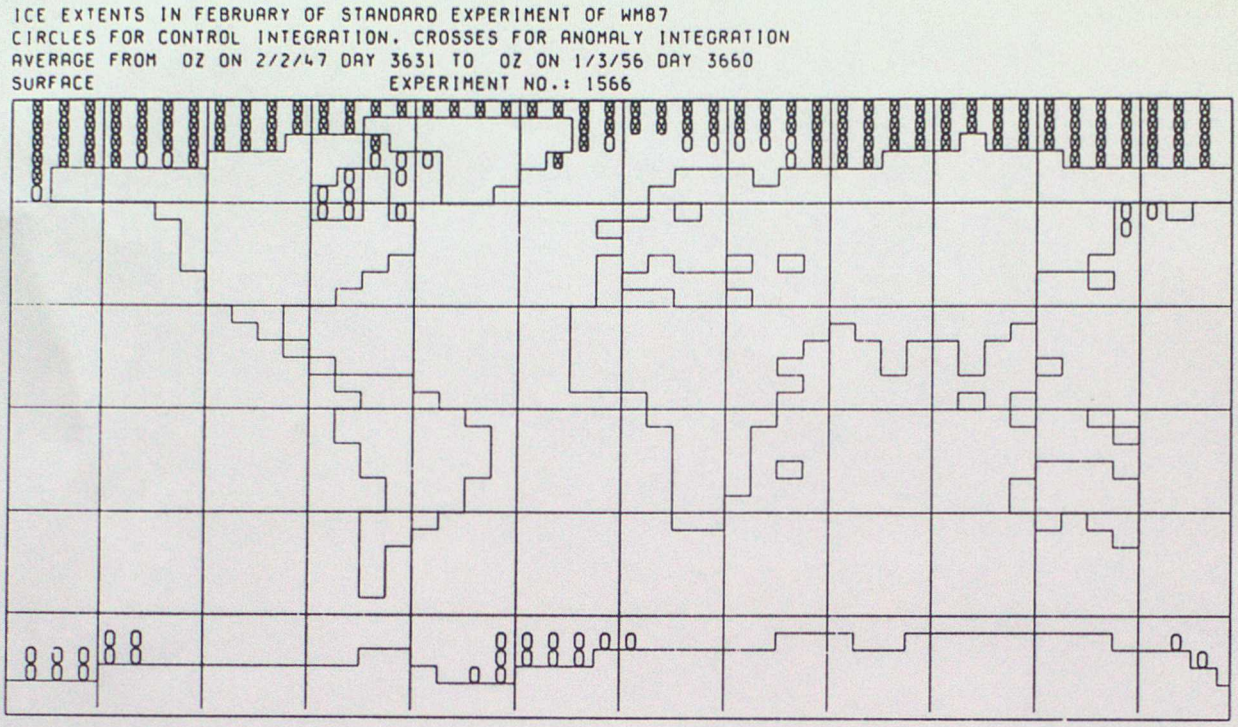


Figure 2. Zonal mean warming due to doubling CO_2 for the standard experiment (dashed line) and the experiment with prescribed sea-ice limits (solid line) meaned over December, January and February (above) and June, July and August (below).



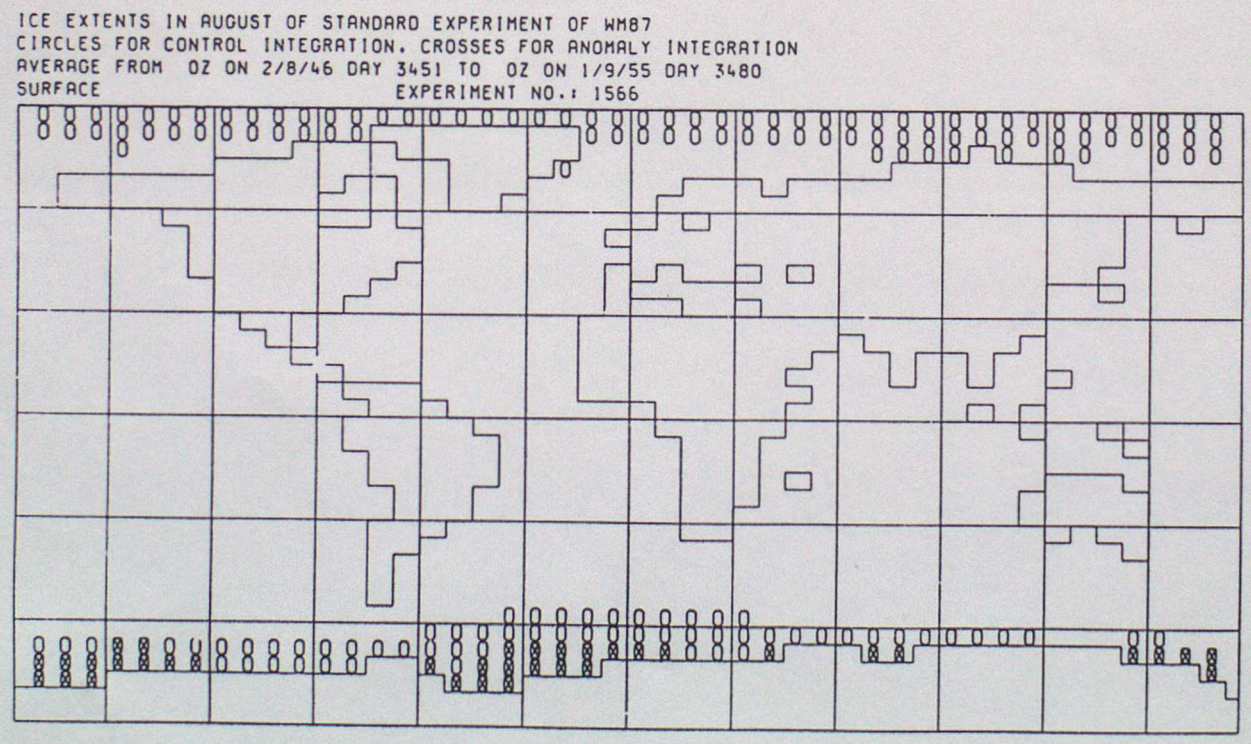
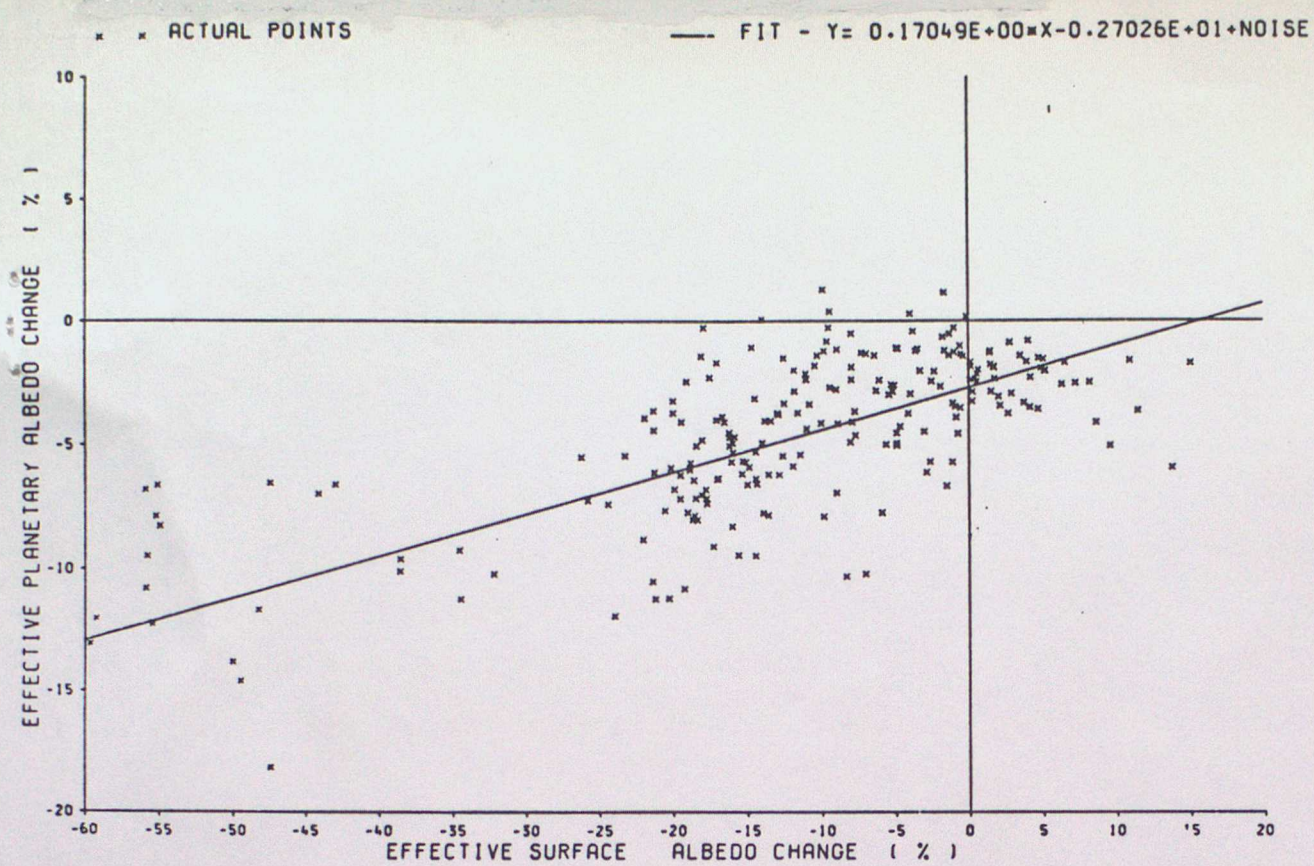


Figure 3. Sea-ice extents in February (above) and August (below) in the control (circles) and doubled CO_2 (crosses) integrations of WM87. Points are marked if they were sea-ice more than half of the time concerned.



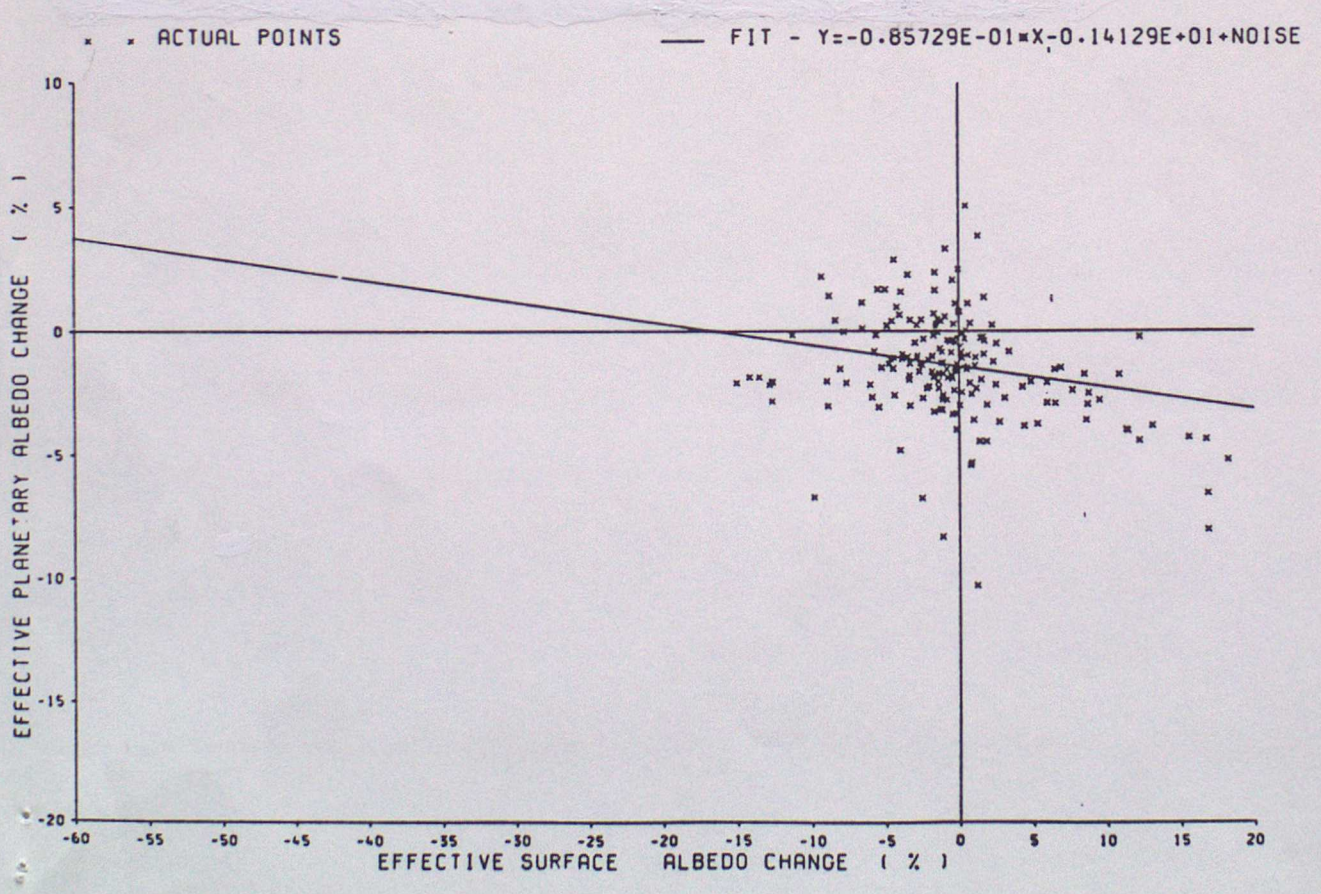


Figure 4. Scatterplots of change at each sea-ice point of the effective surface and planetary albedos, meaned over December, January and February, in the experiment of WM87 (above) and the experiment with prescribed sea-ice limits (below). All points which had sea-ice and received sunlight at any time in any of the four integrations are included. The area-weighted root-mean-square best-fit straight line is also indicated.

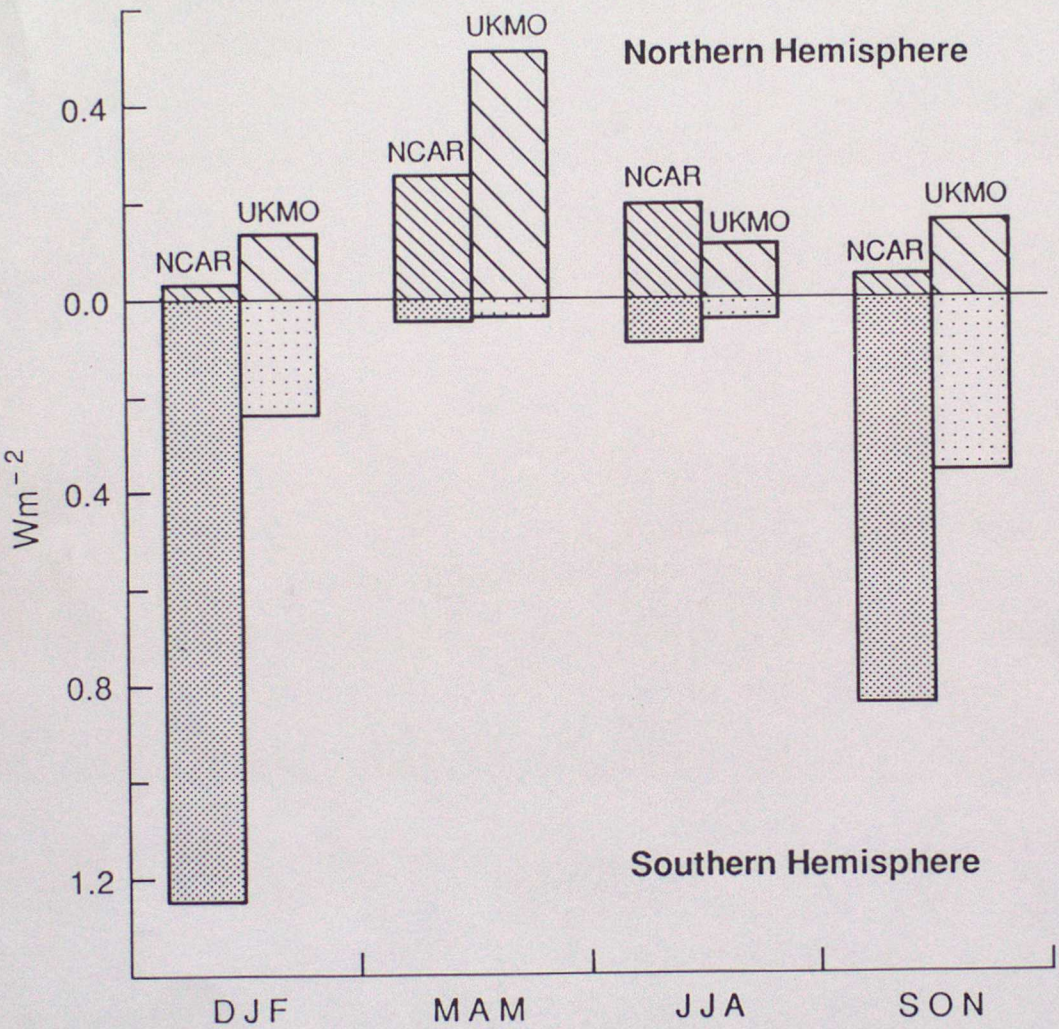


Figure 5. Contribution from each season and each hemisphere to the global annual average decrease in reflected sunlight from the belts where the albedo feedback was estimated to be concentrated in the experiments of Washington and Meehl, 1984, (left-hand column of each pair, reproduced from Dickinson et al., 1987) and WM87 (right-hand columns). Thus these are equivalent to the contributions to the feedback parameter estimated by Method 2a multiplied by the mean surface temperature change. Note that the contributions from each hemisphere are positive away from the zero axis.

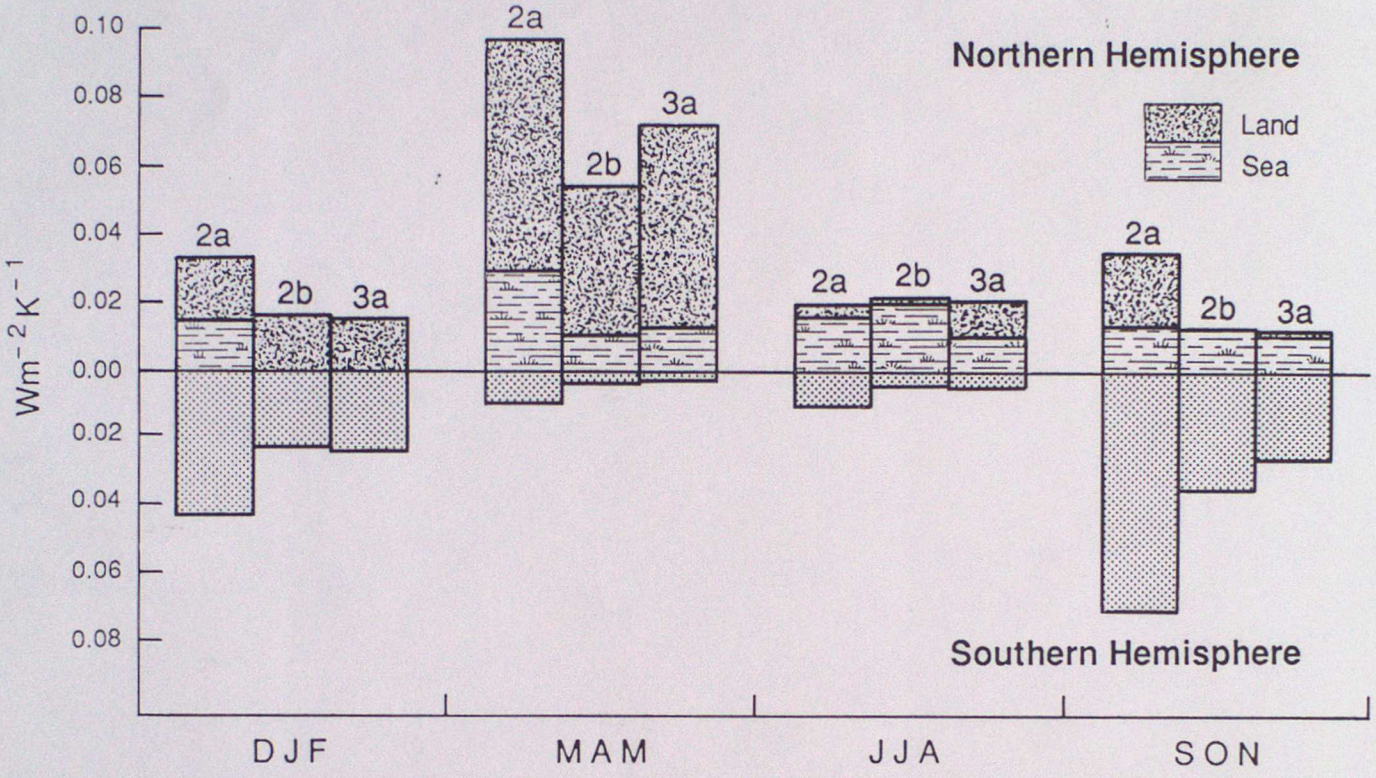


Figure 6. Contributions to the global annual mean feedback parameter in the experiment of WM87 from each season, each hemisphere, and surface type in the Northern hemisphere, for Methods 2a, 2b and 3a (reading from left to right in each group) of Table 4. Note that again the contributions from each hemisphere are positive away from the zero axis.

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